Dynamical Climatology

Surface evaporation during free convection.

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DCTN 80

September 1989

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DYNAMICAL CLIMATOLOGY TECHNICAL NOTE NO. 80

SURFACE EVAPORATION DURING FREE CONVECTION

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1. INTRODUCTION

The largest expanse of warm water in the world occurs in the west Pacific and east Indian Oceans. This warm water is part of the driving centre of the atmospheric general circulation and this circulation is sensitive to even small changes in the sea surface temperatures (SSTs) of this region (Palmer, 1985). Recent coupled model experiments (Gordon, 1989) have also illustrated the sensitivity of the ocean simulation to the net surface heating in the west Pacific. Excessive surface heating in this area leads to a rapid increase in surface temperature which can, due to coupled feedbacks, lead to a propagation of the warm anomaly in the coupled model in a way similar to that observed during recent El Nino events. The heat input into the west Pacific arises from a detailed balance of component heat fluxes. The largest and most variable terms are the short wave and latent heat fluxes. In this note we will be concerned with the parametrization of the latent heat loss in these regions.

The west Pacific is characterized by a warm pool of water with temperatures greater than 28°C. This water is sufficiently warm for large scale convection to occur in association with regions of low level convergence. Winds are generally light, less than 3ms-1 in the climatological mean, over the warmest water. The problem we wish to address here is the parametrization of evaporation in such low wind convectively unstable conditions. There is little oceanic field data for such conditions but there have been a number of studies concerned with evaporation from lakes (Henderson-Sellers, 1986) and in this note parametrizations based on these data will be compared with that currently in use in the Met. Office 11 layer atmospheric general circulation model.

This note is organized as follows: Section 2 reviews the relative importance of wind versus SST changes on variations in surface evaporation. Section 3 describes the parametrization of the free convective limit used in the 11-layer model and Section 4 compares the evaporation rates obtained using this approach with those produced using empirical formulae based on lake data. Section 5 discusses the results and the formulae used.

2. EVAPORATION FROM THE OCEAN

The bulk aerodynamic formula for the latent heat flux is usually written in the form

$$Q=\rho L_E C_E V(q_s(T_S)-q(T_A)) \qquad (2.1)$$

where ρ is the density of air, L_E the latent heat of vaporisation, C_E the exchange coefficient, V the wind speed at the reference level (often 10m), $q_s(T_S)$ the saturation mixing ratio at the SST T_S and $q(T_A)$ the water vapour mixing ratio at the air temperature (T_A) measured at the reference height. The exchange coefficient is often specified as a function of wind speed and atmospheric stability although there is considerable uncertainty regarding its specification and this is certainly the case in the low wind, unstable conditions to be considered here. The simplest approach is to assign C_E a constant value and this appears to be a reasonable approximation for a wide range of atmospheric conditions (Large and Pond, 1982).

It follows from (2.1) that the rate of increase of latent heat loss for unit increase in temperature increases as a function of SST. This is a consequence of the dependence of $q_s(T)$ on T as expressed by the Clausius-Clapeyron equation. This increase with SST is often given as a partial explanation of the importance of tropical (i.e. warm mean temperature) SST anomalies on the general circulation. The effect is illustrated in fig. 1 which shows the latent heat loss as a function of SST for various wind speeds when the air temperature is held fixed at 25°C, the relative humidity at 0.8 and C_E is assigned a constant value of 1.3×10^{-3}. The influence of the wind speed dependence in (2.1) is also illustrated in this diagram. With an SST of 27°C and wind speed of $6ms^{-1}$ the rate of increase of latent heat flux with SST is around $35Wm^{-2}K^{-1}$.

The curves in fig.1 were produced assuming a constant air temperature. Over the ocean the air temperature follows closely, to within a degree or two, the SST and so the assumption of constant air temperature for a wide range of SST is not very realistic. The assumption of constant air temperature would only apply in a situation where the SST suddenly changes and then only for a short period before the atmospheric profile adjusts (rapid changes in air temperature can occur due to coastal effects and on smaller spatial scales due to convective downdraughts). Such sudden changes in SST rarely occur in the open ocean (sudden in this context means: timescale of SST change < atmospheric response timescale) although the effect may be seen in atmospheric model SST anomaly experiments in which an SST anomaly is inserted into a model state in equilibrium with a control SST field. A number of authors have suggested that the assumption that the air-sea temperature difference remains constant over

the tropical oceans is a better approximation than assuming a constant air temperature when parametrizing the surface heat flux (e.g. Philander and Seigel, 1985). Figs. 2a, 2b and 2c show curves for the latent heat flux as a function of SST for various wind speeds assuming a constant air-sea temperature difference of 1°C, 1.5°C and 2°C respectively and a relative humidity of 0.8 (as in the derivation of fig.1). In all the curves in fig.2 the dependence of the latent heat flux on SST is weaker than in fig.1. The rate of increase of latent heating with the SST at 27°C, wind speed 6ms^{-1} and an air-sea temperature difference of 1.5°C is now 8Wm^{-2}K^{-1} compared to 35Wm^{-2}K^{-1} when the air temperature was held fixed. The curves in fig.2 show that the effect of changes in wind speed is now more important than the effect of changes in SST in determining the surface evaporation rate.

In the following sections we will consider in some detail the representation of C_E in situations that are unstable (the air-sea difference will be taken as 1.5°C) and in which the wind speed is low (less than 3ms^{-1}, say). Comparisons will also be made with empirical formulae derived from lake data.

3. THE SURFACE LAYER IN LOW WIND UNSTABLE CONDITIONS

In stable, neutral or mildly unstable conditions Monin-Obukhov similarity theory can be used to parametrize the surface exchanges (Carson, 1976). A brief description of the procedure follows - details can be found in Carson (1976, 1980) and Slingo (1985). The vertical gradient of any quantity (X, say) in the surface layer can be written as

$$\frac{\partial X}{\partial z} = \frac{X_*}{kz}\,\varphi_x(z/L) \qquad (3.1)$$

where X_* is the scaling variable for X (by analogy with the friction velocity u_* for the wind speed U), z is the height coordinate, k is the von Karman constant, L the Monin-Obukhov length and φ_x is a similarity function which may be different for each of the quantities represented by X. The bulk aerodynamic formula for the surface flux of the quantity X is given by

$$F_x = -\rho C_x V \Delta X \qquad (3.2)$$

where ΔX is the difference between the value of X at the reference level and the surface roughness height. An expression for C_x can be obtained by equating

$C_x V\Delta X = u_* X_*$ (this defines X_*) and obtaining ΔX, required in (3.2), by integrating the gradient relationship (3.1) between the roughness height z_x and the reference level.

It is usual in modelling not to work in terms of the Monin-Obukhov length (L in 3.1) directly but rather to eliminate this in favour of the bulk Richardson number defined by

$$Ri_b = \frac{gz}{T_1} \frac{(\theta_1 - \theta_{surf})}{V_1^2} \qquad (3.3)$$

where θ_1 and V_1 are the potential temperature and wind speed respectively at the reference level and T_1 is the temperature at this level. Ri_b is related to the Monin-Obukhov length (L) by

$$z/L = \frac{k C_H Ri_b}{C_D^{3/2}} \qquad (3.4)$$

where C_D and C_H are the momentum and heat exchange coefficients respectively. Equation (3.4) and the expressions for C_x, derived as indicated above, can be solved iteratively for the exchange coefficients as functions of the bulk Richardson number.

In unstable low wind conditions the above analysis breaks down because the friction velocity, and therefore the Monin-Obukhov length, loses its significance as a relevant scaling variable. In the limit of zero wind the dependence of the surface fluxes on the mean wind must clearly vanish. Free convection can still give rise to a flux from the surface and the parametrization of the fluxes in this situation will now be considered. To be specific we will first consider the sensible heat loss from the surface since this is the starting point in the derivation of the lake formulae described later. In zero wind conditions the only parameters that can determine the gradient of potential temperature are the buoyancy parameter g/T (that is, g/T multiplies the potential temperature gradient to give the buoyancy frequency), the surface heat flux $H(0)/\rho c_p$ and the height z (see Carson, 1976 for a detailed argument). Dimensional analysis then gives

$$\frac{\partial \theta}{\partial z} = -c (H(0)/\rho c_p)^{2/3} (g/T)^{-1/3} z^{-4/3} \qquad (3.5)$$

where c is a constant found by experiment to be 0.84 (Carson, 1976). Integration of (3.5) from a height z_1 to a reference height z gives

$$\Delta\theta=\theta(z)-\theta(z_1)= 3c(H(0)/\rho c_p)^{2/3}(g/T)^{-1/3}(1/z^{1/3}-1/z_1^{1/3}) \qquad (3.6)$$

which, after re-arrangement, leads to

$$H(0)/\rho c_p=\alpha(z_1,z)(gz/T)^{1/2}(\Delta\theta)^{3/2} \qquad (3.7)$$

where

$$\alpha(z_1,z)=(3c((z/z_1)^{1/3}-1))^{-3/2} \ . \qquad (3.8)$$

If the heat flux in the free convective limit is written in the bulk form

$$H(0)/\rho c_p=-C_H V\Delta\theta \ . \qquad (3.9)$$

and the wind speed V occurring in (3.9) is replaced by the bulk Richardson number via (3.3), equating (3.7) and (3.9) then gives

$$C_H=\alpha(z_o,z)\sqrt{-Ri_b} \ . \qquad (3.10)$$

This implies C_H α $1/V$ in the limit as V tends to zero. The identification of z_1 with the roughness length z_o in the coefficient α is open to question because in the limit of no wind the roughness length is undefined. Although derived for the sensible heat flux the result (3.10) is used in the 11-level model for both the moisture and heat exchange coefficients and in the next section evaporation rates calculated from it will be compared with those calculated using formulae based on lake data. From the above results the heat flux in zero wind conditions for a reference height z is given by

$$H_o=\rho C_p\alpha(z_o,z)(gz/T)^{1/2}\Delta\theta^{3/2} \ . \qquad (3.11)$$

4. COMPARISON OF EVAPORATION FORMULAE

As noted above, the theory described in the previous section is used as the basis for the current parametrization of the surface exchange coefficients in unstable conditions in the Met. Office 11-level model (Slingo, 1985). In the present context there is a particular uncertainty regarding the conditions in which the free convective limit is reached. In the absence of field observations this can only be specified in an ad hoc way. The parametrization of evaporative loss has been considered in some detail in connection with

evaporative losses from lakes (see Henderson-Sellers, 1986 for a review). In this section we will consider two formulae derived to describe evaporation from lakes and compare these with the evaporation rates produced by the scheme outlined in the previous section. The two formulae are those of Ryan et al (1974) and Sill (1983) which Henderson-Sellers (1986) regarded as the most reliable. Each of these will be described in turn.

4a. Ryan et al (1974) formula

By analogy with convective heat loss from a heated plate Ryan et al (1974) proposed the formula

$$Q=\rho(\lambda(T_s-T_A)^{1/3}+bV_2)(q_s(T_s)-q(T_A))0.662/p \qquad (4.1)$$

where Q is the latent heat flux in Wm^{-2}, p the pressure in mbar, V_2 the 2m wind and λ and b are determined from data to be b = 3.2 $Wm^{-2}mbar^{-1}(ms^{-1})^{-1}$ and λ = 2.7 $Wm^{-2}mbar^{-1}(°C)^{-1/3}$. The term including $\lambda(T_s-T_A)^{1/3}$ represents the contribution from free convection and is assumed to be always present in unstable conditions. Ryan et al (1974) verify this formula against data from cooling ponds in which the size of the pond is relatively small (typically less than $1km^2$) and the water-air temperature differences are considerably larger than those occurring over the ocean. They, however, state that (4.1) is quite general and fits both field and laboratory data well.

4b. Sill (1984) formula

Sill (1984) argues that the simple addition of the free and forced (wind driven) convection contributions in (4.1) is inappropriate. This is because the wind interrupts the convective cells which would form in the absence of wind and in the limit of high winds the free convective contribution to the surface evaporative loss becomes very small. The heat transfer by free convection is via larger eddies than those associated with shear induced transports and the suggestion that the wind shears interact with the convective eddies and reduce the efficiency of the heat transfer is a reasonable one. Sill (1984) suggests the formula

$$Q=\rho L_E C_E V_{10}(1+a_3 C_R)(q_s(T_s)-q(T_A)) \qquad (4.2)$$

where $$C_R=\frac{0.0017}{0.0012V_{10}}(T_s-T_A)^{1/3}$$

and $a_3=0.73C_R$ if $C_R < 1.37$
$a_3=1.0$ $C_R > 1.37$

This result was also tested on lake data. As the wind speed increases the contribution from free convection is proportional to C_R^2, that is, $1/V$ so it vanishes in the high wind limit. Sill (1983) suggests that his result should be applicable to the open ocean.

4c. The 11-level model scheme

The scheme in the 11-level model is based on the theory presented in Section 3. The lowest model σ-level has its centre at approximately 100m above the surface and so this is taken as the reference level. A roughness length of $z_o=10^{-2}$cm is used in the calculation of α in (3.10) and the free convective limit parametrization is used for values of the bulk Richardson number less than -200. For moderately unstable conditions the similarity functions of Dyer and Hicks (1970) are used and these are extrapolated to match the free convective limit form in (3.10) for very unstable conditions.

4d Comparison of results

In order to compare the different formulae conditions in which the air-sea temperature difference is 1.5°C and the SST 30°C were chosen. These are fairly extreme conditions which might be found in the west Pacific warm pool region. Coupled model experiments have indicated that the warmest water is usually located in regions of low winds (Gordon, 1989). In Sill's (1984) formula (4.2) and the standard formula (2.1) the 10m exchange coefficient is taken to be $1.2\text{x}10^{-3}$. The different schemes require the wind speed at different reference levels, ranging from 2m in the Ryan et al (1974) scheme to 100m in the Met. Office model. For simplicity the winds at these different levels were assumed to be related by a logarithmic wind profile with an assumed roughness length of 10^{-2}cm. Potential temperature was assumed well mixed throughout the lowest 100m of the boundary layer.

Fig. 3 shows the calculated latent heat fluxes as a function of wind speed from each of the schemes using the conditions specified above. The Sill curve approaches the standard result for higher wind speeds which is a consequence of

the same value of C_E used in both formulae and illustrates the removal of the free convective contribution at higher speeds. In the Ryan et al formula the free convection contribution is additive to the forced convection part and the curve is therefore a straight line with non-zero intercept. The Met. Office model curve agrees well with Ryan's curve for higher winds although this is fortuitous since free convection is not included in the model scheme at these wind speeds. The differences between the model and standard curve at wind speeds greater than $2ms^{-1}$ probably reflects the different exchange coefficients used. The most significant differences occur at zero wind speed when free convective transports dominate. Both the Sill and Ryan curves indicate values of 35 - 45 $Wm^{-2}K^{-1}$ which are considerably larger heat losses than the model value of around $25Wm^{-2}K^{-1}$. The reasons for this difference will be discussed in the next section. It is interesting to note that for the above conditions the model scheme does not come into effect until the wind drops to below 0.1 ms^{-1} (once the free convective limit (3.10) is used the model flux becomes independent of wind speed). This very low wind arises as a consequence of the -200 limit placed on the bulk Richardson number and is probably rarely reached in the GCM. Over land the GCM assumes a limit of -5 on Ri_b before the free convective limit comes into play and this gives more reasonable values for the limiting winds of approximately $1ms^{-1}$.

5. DISCUSSION

If the Ryan and Sill formulae are written in the bulk form given by (3.2) then in the free convective limit

$$C_E = 0.0017 \frac{(T_s - T_a)^{1/3}}{V_{10}} \quad - \text{RYAN} \qquad (5.1)$$

$$C_E = \lambda (p/0.662) \frac{(T_s - T_a)^{1/3}}{V_{10}} \quad - \text{SILL} \qquad (5.2)$$

$$C_E = \alpha(z_o, 100)(100g/T_1)^{1/2} \frac{(\theta_1 - \theta_{surf})^{1/2}}{V_{10}} \quad - \text{MODEL} \qquad (5.3)$$

The model result arises from the assumption that $C_E = C_H$ and the relations (3.10) and (3.3). The essential difference between the lake and the model formulae occurs in the different functional dependence on the air-sea temperature difference. This arises because molecular effects are included in the lake formulae but not in the model. The limiting form of the lake formulae are

derived using dimensional arguments that show (see Turner, 1979) that the ratio of the free convective heat transport to the molecular diffusive heat transport (the 'Nusselt number') must be a function of the Rayleigh number. General arguments show that this ratio must be a function of $Ra^{1/3}$ where Ra is the Rayleigh number (see Turner, 1979 page 213). The physical processes inhibiting the convection and mixing with the environment are assumed to be molecular in origin. The coefficients in the formula are determined by laboratory experiments where there is no wind. The applicability of these formulae to the open ocean, as claimed by Sill, is questionable since there is generally a background wind, which although small, generates more mixing than molecular effects (see Priestly, 1959, page 66).

Despite this limitation it can be assumed that the free convective heat flux arising from convective parcels in which environmental mixing is molecular in origin (rather than turbulent) should be regarded as a minimum on the free convective heat flux. This approach was used by Deardorff (1972) who took Townsend's (1964) form of the free convective flux (which is essentially equivalent to the lake formulae at zero wind) as a minimum flux during free convection. Townsend's results were derived from a laboratory tank experiment.

There are significant differences between the evaporation rates predicted by the lake formulae and those produced by the model scheme and it is of particular concern that the model values are considerably smaller. To increase the model evaporation rates due to free convection would require an increase in the value of α appearing in (5.3). The value currently used in the 11-layer model scheme is obtained from (3.8) with $z_1 = z_o = 10^{-2}$cm, the roughness length assumed over the ocean. Considering that z_o is not defined in zero wind conditions and that z_o is also a function of wind speed the use of (3.8) to determine α must be regarded with caution. If a pragmatic approach is taken in which the limiting bulk Richardson number is changed to -5 (the value over land) and α is chosen so that the free convective limit values of the exchange coefficients match the existing values at this limiting Ri_b, the curve in figure 4 is obtained. This shows considerably better agreement with the lake formulae. An alternative approach would be to follow Deardorff (1972) and use the evaporation limited by Townsend's (1964) flux as a minimum value.

The differences of around 25 Wm^{-2} at zero wind between the different formulae may seem rather small but in these conditions the ocean mixed layer will be shallow and therefore SST is very sensitive to changes in heat input. This has

been a particularly important consideration in recent coupled model simulations (Gordon, 1989).

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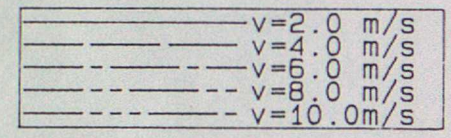


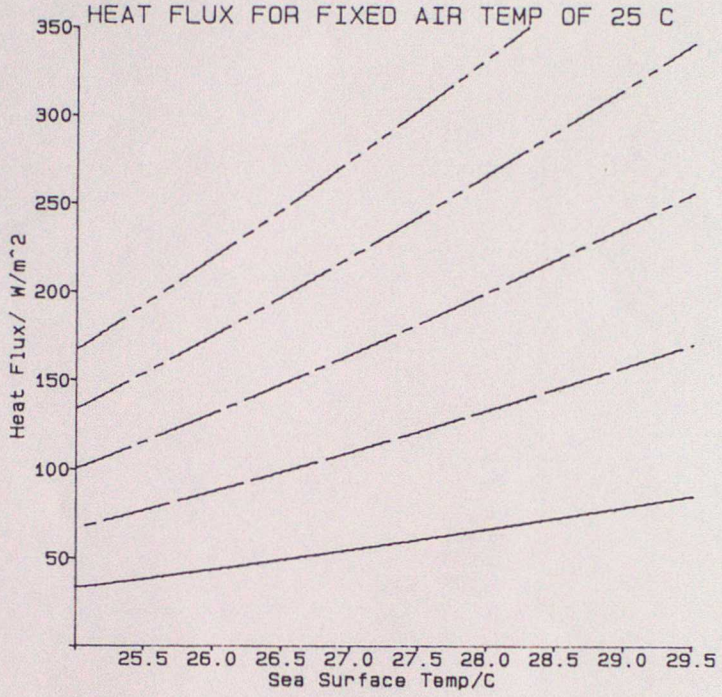
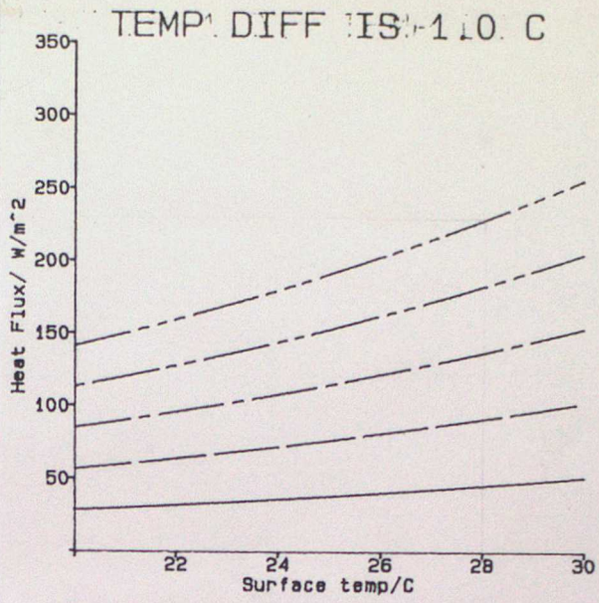
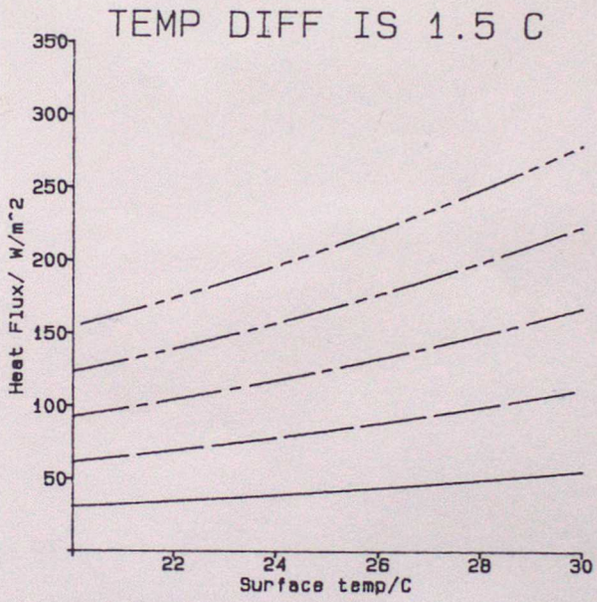


Fig.1. : The latent heat loss from the surface as a function of SST for various wind speeds when the air temperature is held fixed at 25°C and C_E is assigned a constant value of 1.3×10^{-3}.

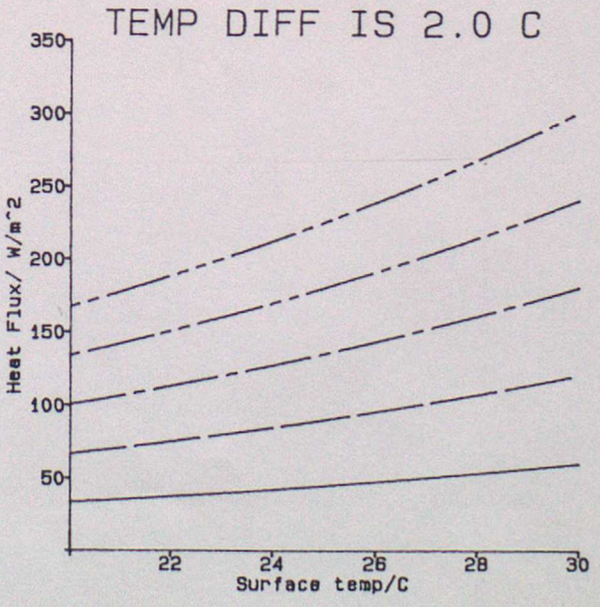
Fig.2. :(a) The latent heat loss from the surface as a function of SST for various wind speeds when the air temperature is taken to be 1°C colder than the SST and C_E is assigned a constant value of 1.3×10^{-3}.

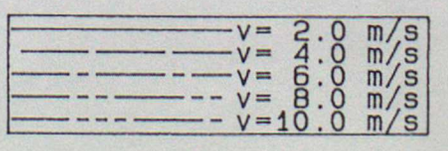


(b) As fig.2a but with an air–sea temperature difference of 1.5°C.



(c) As fig.2a but with an air–sea temperature difference of 2.0°C.





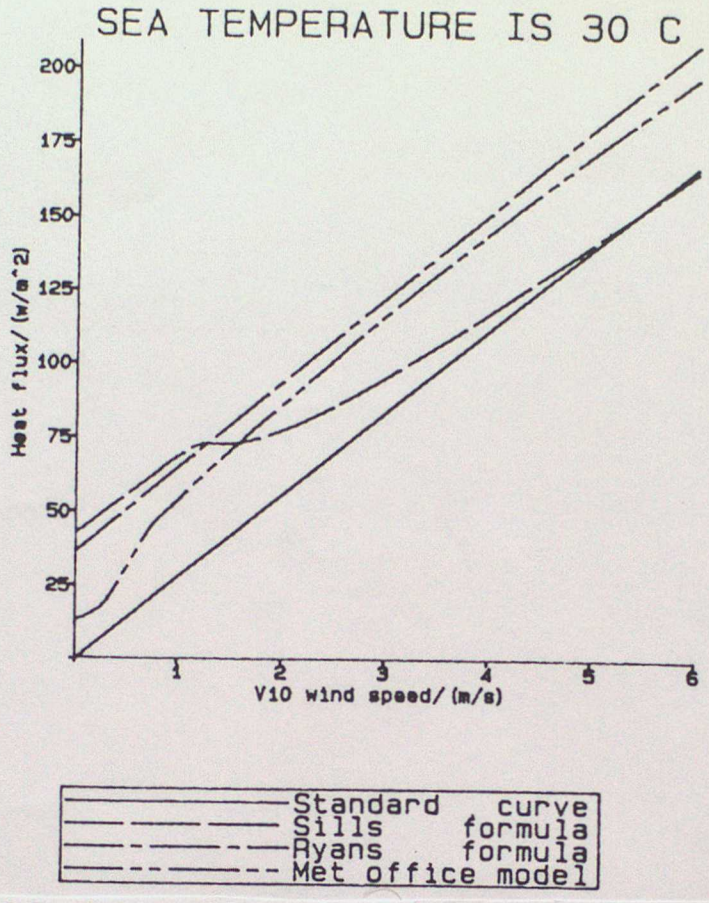


Fig.3. : The latent heat loss from the surface as a function of wind speed as calculated using the standard formula with constant C_E given in (2.1) and the schemes of Ryan et al (1974), Sill (1983) and that used in the 11-level model. The SST is held fixed at 30°C, the air-sea temperature difference is 1.5°C and the relative humidity is 0.8.

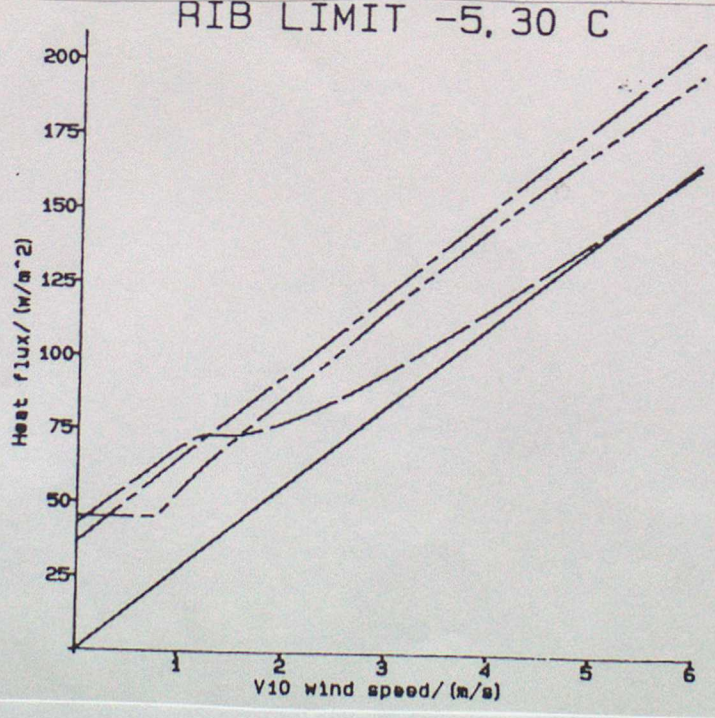


Fig.4. : As fig.3 but with the free convection parametrization of the exchange coefficient switched on at Rib=-5 rather than -200 and the value of α chosen to ensure continuity with the value of the coefficient for smaller Rib.

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DCTN 1	June 1984	A scheme for incorporating the thermodynamic sea-ice model into a coupled ocean-atmosphere model. C Cordon and M Bottomley
DCTN 2	Sept 1984	The parametrization of the upper ocean mixed layer in coupled ocean/atmosphere models. C Cordon and M Bottomley
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